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Snow cover duration and extent for Great Britain in a changing climate: altitudinal variations and synoptic-scale influences

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Abstract: Snow cover is an important indicator of climate change but constraints on observational data quality can limit interpretation of spatial and temporal variability, especially in mountain areas. This issue was addressed using archived data from the Snow Survey of Great Britain to infer key climate relationships which were then used to reference larger-scale patterns of change. Data analysis using non-linear (logistic) regression showed average changes in yearly snow cover were strongly related to mean temperature rather than precipitation values. Inferred change shows long-term decline in average yearly snow cover with greatest declines in some mountain areas, notably in northern England, that can be related to their position on the most temperature-sensitive segment of the logistic curve. Further declines in snow cover were projected in the future: a central ensemble projection from HadRM3 climate model showed average yearly snow cover predominantly confined to Great Britain mountain areas by the 2050s. However, interannual variability means some years can deviate significantly from average snow cover patterns. Site-based analysis showed this variability has distinctive geographical variations and different influences for mountains compared to adjacent valleys. Comparison of inter-annual variability with Lamb weather type frequency and NAO index shows the influence of large-scale airflow patterns on snow cover duration. Most notable is the role of northwesterly and northerly flows in explaining snowy years on mountains exposed to that direction, compared to influence of easterly flows at lower levels. Future changes will therefore depend on dominant annual/decadal circulation patterns in addition to long-term declines from climate warming.

Keywords: snow cover, climate change, Lamb weather types, logistic regression, mountains, synoptic-scale, Great Britain

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1. Introduction

Snow cover acts as a modifying interface between atmosphere and ground, altering surface albedo, thermal insulation properties, hydrological cycle, and biosphere processes (Cohen 1994). In temperate latitudes, effects are most pronounced in mountain regions where snow cover is more frequent (Barry 1992; Beniston 2006). By contrast, temperate lowlands have more ephemeral snow cover but nevertheless with potential to disrupt human activity. Following periods of snowfall, meteorological conditions favouring persistent snow cover are continuation of lower temperatures that delay snow melt, possibly accompanied by freeze-thaw cycles that consolidate the snowpack (Dunn et al. 2001). Variation in synoptic climatology means that patterns of snow accumulation and snow cover can vary strongly by location and from year to year, particularly in temperate oceanic regions (Stewart 2009). This climatological sensitivity highlights the potential combined value of snow cover duration and extent as key indicators of climate change, especially in locations with ephemeral snow where maximum snow extent is a less useful indicator. Nevertheless, significant challenges remain in analysing spatiotemporal variations in snow cover, primarily due to limitations of existing observational data, especially in mountain regions. Snow extent and duration also vary at a local scale due to topographic influence on snowfall rates, wind exposure, solar exposure, and temperature. Hence, model-based simulations of changes in snow cover have often had difficulty when validating models against observed variability (past and present), which constrains the confidence in future projections, as has occurred for Great Britain (GB; Bell et al. 2016). To address these challenges, the present study developed an approach using archived historical data to derive generalised relationships between synoptic-scale climatology and snow cover that can be then used to contextualise recent and future change.

Changes in duration and extent of snow cover have important implications. Snow albedo and snow melt influence local climatology through energy fluxes (Pomeroy and Brun 2001), and the duration of water storage in the snowpack affects soil moisture and catchment hydrology (Dunn et al. 2001; Diffenbaugh et al. 2013). Snow cover can increase the likelihood of extreme minimum air temperatures close to the earth surface whilst acting as ground insulator, moderating soil temperatures and protecting plants from frost damage (Oke 1987). Snow presence therefore affects differential survival rates of plant species and hence distribution of vegetation communities, with snow-melt timing representing a key stage in growth cycles of arctic-alpine plants (Keller et al. 2005; Cutler 2011; Gottfried et al. 2011). Snow cover protects the ground from subaerial weathering and erosion associated with freeze-thaw and water runoff, including possible additional damage from human or animal trampling (Schlochtern et al. 2014). However, snow cover may also concentrate deposition of atmospheric pollutants with consequent impacts on soil, plant, and aquatic communities (Dore et al. 1992; Helliwell et al. 1998).

Socio-economic implications of snow cover variability are directly evident in locations dependent on winter sports activities, or where disruption to transport and other infrastructure occurs (Harrison et al. 2001). Farming may experience negative effects through snow-related disruption to land management and food supply problems for livestock, especially in marginal upland areas (Jones et al. 2012). Impacts may also be severe during anomalous snowy years at locations that normally have limited snow cover and where precautions are not usually considered necessary. Beyond its direct influence, variations in snow cover can have high social and cultural relevance: snow-covered landscapes have been shown to have an important role in individual and collective memory of the weather and in local landscape identity (Manley 1952; Hall and Endfield 2016).

Studies have generally reported declining Northern Hemisphere snow cover duration where a clear trend exists (e.g. Brown 2000; Durand et al. 2009; Choi et al. 2010; Marty et al. 2017; Beniston et al. 2018). Negative trends are predominantly from lower elevations, suggesting threshold effects exist related to the combined effect of temperature and precipitation on snowfall quantities as they co-vary across both altitude (based upon the lapse rate) and latitude/longitude (Stewart 2009; Fontrodona Bach et al. 2018). Such threshold effects may evolve further as future modelling suggest anthropogenic climate change will bring both warming and additional winter precipitation for many northern temperate regions (Brown and Mote 2008; Kay 2016). Long-term changes due to external forcing will be modulated through influence of internal shorter-term variations in atmospheric circulation on temperature and precipitation patterns, as for example occurring between different modes of the North Atlantic Oscillation (Stewart 2009; Irannezhad et al. 2016).

Sensitive snow-cover regions have been identified as having seasonal mean air temperatures in the range -6°C to +6°C, notably midlatitude coastal margins and associated mountain ranges (Brown and Mote 2009; Adam et al. 2009). Although GB occupies this transitional zone, knowledge of recent changes in snow cover is mainly confined to specific locations or individual years: Harrison (1993) analysed different patterns of snow cover in Scotland using indicative 'cold' and 'mild' winters, whilst Trivedi et al. (2007) used site-level analysis at Ben Lawers (Figure 1) to infer that snow cover at higher altitudes may be more sensitive to climate variability than previously thought. This highlights requirements for larger-scale analysis that can facilitate interpretation of long-term change (Watts et al. 2015) and validation of simulation models (Bell et al. 2016).

Process-based models using energy-balance schemes can simulate snow accumulation and melt but require detailed parameterization for multiple variables, meaning they are most suited to small-scale study sites (e.g. Uhlmann et al. 2009; Essery et al. 2013). Energy-balance equations embedded within global or regional climate models are necessarily resolved at a coarse resolution (Räisanen 2008), hence parameterization is not intended to simulate local spatial variability of snow cover as required for mountain areas (Keller et al. 2005; Dutra et al. 2011). These challenges have encouraged the use of reduced complexity, large-scale approaches based upon empirical association of snow cover with finer-resolution temperature and precipitation data, generally considered adequate 'for most practical purposes' (Ohmura 2001). Temperature-index methods (also including degree-day models), as commonly used in snowmelt hydrology, define snow storage conditions according to a threshold temperature supplemented by other factors (Hock 2003). For example, the snow module of Bell et al. (2016) simulated snow cover based upon temperature and precipitation transition equations in order to model hydrological variability due to climate change in GB. By contrast, statistical methods infer such relations from observation data. Temperature data in the Alps have been used empirically to evaluate the sensitivity of snow cover to recent and future warming (Hantel et al. 2000; Hantel and Hirtl-Wielke 2007). Similarly, joint temperature/precipitation distributions have been used as proxies for frequency of synoptic weather patterns and thus to infer relative snowiness of winters in the Swiss Alps (e.g. Beniston et al. 2011), including altitudinal threshold effects (Morán-Tejeda et al. 2013).

Snow data can be accessed from remote sensing sources in addition to field observations (Linde and Grab 2011). However, earlier satellite records dating back to the 1960s are of course resolution and typically contain greater inaccuracies as techniques were refined, which can hinder time-series assessment. Difficulties in interpretation of remote sensing data can result in inconsistencies compared to ground observations, especially due to scale effects and in mountainous areas (Nolin 2010). Cloud cover is often the most severe problem for visible spectra due to similar reflectance properties to snow, but other notable issues include shadows and the presence of forests and bare rocks (Dietz et al. 2012a). Comparison of MODIS satellite reflectance data with ground observation data in Scotland has shown large differences in interpreted snow cover with distinctive biases due to varying effects of cloud cover (Spencer and Essery 2016). Microwave satellite data have some advantages over optical imagery but may produce significant underestimation of snow cover in early and late season when snow is thin or wet, conditions often found in GB (Kelly, 2000; Rees and Steel 2001; Butt, 2006).

Due to the foregoing issues, and challenges in calibrating temperature-index methods for mountain areas (Hock 2003), the present study adopted an empirical statistical approach based upon archived ground survey data of GB snow cover to identify climatological relationships with temperature and precipitation. This procedure could accommodate non-linear relations. Derived relationships were then used to investigate how duration and extent of snow cover have varied in recent decades and may vary into the future. The scope of investigation included both changes in long-term averages of snow cover together with differences that may occur from year to year, as potentially influenced by variability in synoptic-scale atmospheric circulation. An additional objective was to identify any differences that may have occurred at higher elevations in mountain areas compared to the lower elevations from which most inferences of change have been derived, and hence the relative sensitivity of snow cover at different altitudes and locations.

2. Data and Methods

2.1 Data sources

Two main sources of snow data were investigated regarding the research objectives. Firstly, UK Met Office (UKMO) station archives, which provide standard sources for weather observations. Stations recorded a "snow lying" day if at 0900 UTC the countryside at the same level and typical of the station itself was more than 50% snow-covered. Observations have a subjective element and may be biased towards specific aspects but discrepancies can be minimised using multiple stations. UKMO station data have previously been used to derive average snow cover duration maps using inferred altitudinal relations (Manley 1939; Jackson 1978). More recently, automated spatial interpolation (linear regression and inverse-distance weighting of residuals) have enabled a UKMO 5km gridded climatology (1961-2011) generated from UK station data, including snow cover derived from relationships with elevation and terrain shape (Perry and Hollis 2005). Although UKMO gridded data have been used to infer declining trends in snow cover (Biggs and Atkinson 2011; Kay 2016), there are very few meteorological stations above 300m meaning snow cover for the uplands was primarily interpolated from lowland stations. This interpolation is likely to produce substantial errors, particularly for the frequent occasions when lowlands are snow-free but the uplands retain a snow cover. Increased use of automatic weather stations has also reduced observations of surrounding snow cover.

Another data source is the Snow Survey of Great Britain (SSGB) which operated from 1937-2007 (Met Office 2018). SSGB monitored the snowline (the typical elevation where snow covered more than 50% ground) using voluntary observations, with emphasis on upland regions, usually from October to May (Spencer et al. 2014). Volunteers represented diverse organisations including estates, water authorities, conservation agencies, energy companies, and forestry. At each SSGB base station, observers noted the snowline elevation to the nearest 150m up to the highest visible summits. If cloud obscured observations this was recorded; some observers would also interpolate the snowline for these days from adjacent days (cf. Trivedi et al. 2007). Again, observations contain a subjective component which can produce anomalous records for individual stations. However, annual UKMO SSGB summary reports produced for 1953-1992 (Met Office 2018) included a form of quality control whereby snow profiles for specific mountain ranges from individual base stations were supplemented by adjacent stations in the case of anomalous data. In 1994 participating stations were reviewed and observers were no longer required to note null observations. This limits the utility of the post-1993 records as it is often not possible to distinguish absence of snow cover from null observations.

The present study utilised historical SSGB data to better represent snow cover patterns at all altitudes. Comparison of UKMO 5km gridded data and SSGB station data has shown that the UKMO data significantly underestimate snow cover at higher elevations (Spencer et al. 2014). In addition, the relative coarseness of the UKMO 5km grid means that fine-scale variations in snow cover, particularly in the mountains, cannot be adequately analysed (Spencer and Essery 2016).

Following previous studies (e.g. Harrison et al. 2001; Trivedi et al. 2007), snow cover duration (S_d) was defined as number of snow cover days (not necessarily continuous) from October-May; no depth distinction was available to define ‘snow cover’ as adopted elsewhere (e.g. Hantel et al. 2000). Snow cover often persists in isolated patches on GB mountains outside October-May, but patch distribution is strongly related to local topographic influences favouring both snow accumulation through local wind-fields and minimal snowmelt due to insolation shade (Dunn et al. 2001). Analysis of such patches is therefore more suited to a different mode of analysis (e.g. Watson et al. 2002).

2.2 Data preparation

Analysis was primarily based on the compiled mountain-range observation data provided in the annual SSGB reports for 1960/61-1991/92, covering locations from the Brecon Beacons, Snowdonia, Cumbrian Fells, North Pennines, Cheviots, Southern Uplands and throughout the Scottish Highlands (Figure 1). These reports provide summary monthly and annual (October-May) S_d data at station level, at 750m, and close to summit level for named peaks. With regard to missing data, if a single month was missing from an annual record then it was estimated by linear interpolation using comparison of complete monthly records from adjacent stations, otherwise that location was excluded from analysis. This provided 349 yearly S_d records at 3 altitudes (hence n = 1047). SSGB locations varied through time and snowline observations of the same mountain range for different years may therefore be derived from different stations, which can make direct comparisons through time difficult. Hence, although all available SSGB mountain-range annual data were used for the GB scale assessment, subsequent investigation of local S_d variability was based upon selected SSGB locations that consistently provided observations throughout the 1960-1992 analysis period. Local variability analysis was also facilitated by raw digital SSGB data for some locations in Scotland (Spencer 2016b) when complementarity with the SSGB report data allowed completion of the time series.

Temperature and precipitation data were also collated for the SSGB locations described above. Although some SSGB locations were equivalent to UKMO stations, potentially allowing use of archived primary meteorological data, this was most commonly not the case. Hence, consistent with the large-scale mode of analysis, data were derived from a single source, the UKMO 5km gridded climatology referred to earlier (Perry and Hollis 2005), rather than attempting to reference values at SSGB locations based upon nearest available UKMO stations. Daily mean temperature (T_m) data were processed to derive annual (October-May) average means for each relevant SSGB location based upon associated UKMO climatology grid cell values. T_m data were adjusted to match the appropriate altitude of the SSGB record (i.e. station altitude, 750m altitude, or summit altitude) according to a standard lapse rate adjustment of 0.6°C/100m (Vuille 2014) applied relative to the UKMO 5km grid cell altitude. Although lapse rates can vary in both time and space (Pepin 2001), the broad-scale of assessment and use of yearly summary data support use of a universal constant rate as a simplifying assumption.

A variation was used to derive precipitation data for SSGB locations. Data from UKMO stations and associated gridded data represent total precipitation (P_t: both solid and liquid phases). Although it would be expected that increased snow cover would be expected to be positively correlated with increased snowfall precipitation, this correlation may be reversed if increased precipitation amounts represented rainfall instead of snowfall, especially as rain can induce rapid snow melt (e.g. Cohen et al. 2015). Hence, in addition to use of P_t values derived from relevant cells of the UKMO 5km gridded data, an additional estimation of snowfall precipitation amount (P_s) was tentatively derived. Precipitation phases can be influenced by factors other than air temperature meaning snowfall can occur at a wider range of temperatures than those close to 0°C (Serquet et al. 2013; Feiccabrino et al. 2015). Simple techniques to apportion solid and liquid precipitation therefore use a transition relationship between lower and upper temperature limits when 100% snow or 100% rain would occur (Anderson 1973; Quick and Pipes 1976). Large-scale assessments suggest that a 50:50 snow-rain ratio occurs at ca. 2°C for humid temperate climates of NW Europe (Jennings et al. 2018); therefore this value was used in the present study with lower and upper limits for the transition range set at -4°C to 8°C mean temperatures (October to May). Unlike T_m, further processing of the UKMO 5km gridded precipitation data to allow for local orographic effects not fully incorporated in the source data was not applied due to lack of supporting evidence. Use of standard precipitation uplift factors for elevation seem inappropriate for the heterogeneous GB climate (Brunsdon et al. 2007; Brown et al. 2017). For P_s, uncertainty on the role of orography is exacerbated by both limited data from upland sites and challenges for rain gauges in capturing snowfall (Dore et al. 1992; Rasmussen et al. 2012).

2.3 Analysis of climate relationships

Regression analysis was used to investigate relationships between S_d and equivalent T_m and P_s data. Analysis was initiated through linear regression but further developed through logistic regression to accommodate non-linear relations through logit functions (Hosmer et al. 2013). There are naturally lower and upper bounds to S_d (0 days and 243 days for full cover October-May respectively, ignoring leap years), whilst coastal influences have an increasing effect at the lower end of the S_d range, suggesting non-linear climate relations (Harrison 1993). Logistic regression was based upon proportional S_d (from 0 to 1.0, with 243 days representing total cover of 1.0) with the aim of finding a good model relationship consistent with the underlying climatology. Iterative exploration of logit values was used to derive suitable values for the intercept, T_m and P_s coefficients by maximising log-likelihood values. Significance testing of T_m and P_s coefficients employed the likelihood-ratio test which generally performs better than other equivalent measures (e.g. Wald test; cf. Jennings 1986).

The finalised regression relationship was used to map GB spatial variability of S_d using source data (T_m or P_s) from the UKMO 5km gridded climatology. To enable finer resolution detail consistent with local topography, T_m data were further downscaled to a 500m grid by using the standard temperature lapse rate referred to above (0.6°C/100m) but no downscaling was applied to P_s as also noted above. To reference recent long-term changes, the periods 1960-1990, 1970-2000 and 1980-2010, were used for analysis. In addition, to investigate future changing patterns of snow cover, climate data were derived from the HadRM3 25km climate model used to derive the UK Climate Projections 2009 (UKCP09; Murphy et al. 2009). HadRM3 data were available as a 11-member Perturbed Physics Ensemble (PPE) which featured variation of multiple parameter combinations to systematically represent key model uncertainties in each ensemble member. Ensemble members could therefore be used to derive projected future change factors for temperature and precipitation, referenced to the standard (1961-1990) baseline, which were then added to the UKMO climatology data to provide future values with simple adjustment (delta-change method) to remove significant biases. To present an indicative central projection for future changes expected by the 2050s, the PPE ensemble mean was calculated for the IPCC A1B scenario (broadly equivalent to a Representative Concentration Pathway of 6.0Wm^{-2} radiative forcing: van Vuuren et al., 2011). Climate data were then used in the regression equation for snow cover. In producing maps, no adjustment was made for land surface effects on snow cover (e.g. lakes or forests).

Further analysis was conducted at 8 SSGB paired locations consisting of mountain peaks (Figure 1) and valley stations for which consistent quality-controlled time-series S_d records could be obtained for 1960-1992; each pair had 28-32 years data available for this period. Correlation analysis (Pearson coefficients) was employed to investigate local variations in S_d at valley and summit altitudes relative to T_m, P_t and P_s derived for those locations from the UKMO gridded climatology. In addition, correlation testing investigated the role of synoptic-scale circulation patterns in yearly S_d variability based upon assumed seasonal synoptic relations with temperature and precipitation values. Analysis included the North Atlantic Oscillation (NAO), which has featured in previous analysis of inter-annual S_d (e.g. Spencer and Essery 2016), using the December-March NAO index of Hurrell et al. (2003). Large-scale climatological influences were also explored through frequency of Lamb Weather Types (LWTs) over the same October-May period as for S_d. LWTs summarise on a daily basis the dominant circulation pattern according to a standard typology derived from geostrophic flow strength, direction, and vorticity (Lamb 1972), routinely calculated and archived following an objective procedure (Jenkinson and Collison 1977). LWT data were sourced from the CRU archive, which derives LWTs from reanalysis data with a grid centred on the UK (Jones et al. 2013). Annual frequency during the October-May period of the 7 seven principal LWTs (anticyclonic A; cyclonic C; northerly N; easterly E; southerly S; westerly W; northwesterly NW) was calculated from the original 27 LWT classes by reallocating hybrid classes in frequency counts (e.g. AS one half to both A and S; CNE one-third to each of C, N and E). It should be noted that Lamb (1972) identified NW as a principle LWT distinct from W and N and not a hybrid, and this distinction was maintained.

3 Results

An initial linear regression model showed a very strong relationship of S_d with T_m (r^2 = 0.68; p <0.001) but no significant relationship with P_t or P_s. A strong positive relationship between S_d and P_s was found with a data subset from base station altitude only but T_m and P_s show strong collinearity at this lower elevation level (Pearson's r = -0.81), and as T_m was used in derivation of P_s the dominant relationship is clearly through T_m. Linear regression was evidently less suitable for fitting upper and lower ranges of S_d data, as shown for the T_m relationship in Figure 2, most notably where higher T_m values would predict a negative S_d (rather than 0 days). A sigmoidal curve derived from logistic regression therefore provides a more logical solution to infer S_d. Following this rationale, logit values for intercept, T_m and P_s were tested with emphasis on ensuring the logistic curve was appropriate across the full range of S_d. A null hypothesis that regression coefficients were 0 was rejected for T_m ($p < 0.001$) but retained for P_s, which is consistent with findings from linear regression. The final logistic function for S_d was therefore simplified and based on T_m only (i.e. logit(S_d) = 0.6 - 0.615T_m); the resulting sigmoidal curve (Figure 2) featuring a mid-range segment of steeper gradient indicating increased sensitivity of snow cover to incremental temperature changes within this range.

Analysis also suggests that a S_d logistic function based on T_m is relatively invariant with time: when the data were disaggregated and analysed over two separate time periods (1960-1975; 1976-1992), very similar logistic functions were derived for both periods based upon maximised likelihood values. Objective validation of the logistic model is not possible because station observation data include subjective components. However, a comparison was made for average (mean) 1960-1980 S_d values calculated from the T_m logistic function against observed snow cover from the 8 SSGB stations with long time-series and with published UKMO station data (Stirling 1997). Making allowances for uncertainties regarding station 'surrounding area', this suggested that the regression-derived values appeared to provide a reasonably close correspondence (within ca.±5%) to SSGB observed S_d on mountain summits, with the exception of some western coastal mountains (Corserine and Cader Idris: Figure 1) where there seemed to be a possible overestimate by 20-25%. For mountain valley or lowland locations, data correspondence also appeared reasonable, although some station data suggest logistic function values may underestimate observed snow cover by 5-10% at those locations. However, 1960-1980 station S_d mean values were often inflated by large values from anomalous snowy years which may account for such discrepancies, with median values being closer to the logistic values in these cases. In addition, differences may stem from the observed data being based only upon snow cover at 0900 UTC omitting the possibility that snow may subsequently melt during the day. Lower S_d values derived from the logistic function may therefore better indicate a full day of snow cover without snowmelt, especially for locations that occupy the higher end of the T_m range.

The logistic function was used to derive GB period maps for average yearly S_d (Figure 3). Maps show a general northerly and easterly increase in S_d with largest values in mountain ranges, together with S_d differences between mountains and valleys or lowland areas within the same region. The snowiest area is confirmed to be the Cairngorms (NE Scotland) where mean S_d is modelled at its maximum (212 days: 1960-1990; 205 days: 1980-2010). Maps also show districts to the south or west of mountain ranges with reduced S_d, as identified from previous time-periods (Manley 1939; Jackson 1978); exclusive use of T_m in the logistic function implies attribution to warmer temperatures in those areas. A general decline in S_d is shown between 1960-1990 and 1980-2010 (Figure 3d), with the largest decreases of more than 20 days, as particularly exemplified by uplands of north England (North Pennines and Cumbrian Fells).

When using the same logistic function for future changes, average yearly S_d was shown to decline at a magnitude greater than recent changes under a central ensemble projection for the 2050s (Figure 4; Table 1). In this model projection, S_d longer than a few days becomes confined to northern and upland areas and only the highest mountains have average durations in excess of 100 days (Cairngorms maximum value: 153 days). By comparison to this central estimate, other future climate projections will imply more or less changes in snow cover, but consistent with scientific consensus on future temperature rises, all HadRM3 PPE model runs show increased temperatures for the 2050s, and hence all indications are for substantial reduction in average yearly S_d.

Time-series analysis at long-record SSGB locations allowed further evaluation of geographical and altitudinal S_d relationships. Analysis confirmed strong correlations with T_m at both valley and summit altitudes and no clear correlation with P_t (Table 2). Positive correlations between P_s and S_d were shown to be strongly positive at all valley level locations but as noted earlier P_s and T_m at this level are themselves strongly correlated and the dominant relationship is through T_m.

Further information on P_s variability at these SSGB locations was obtained from evaluation against synoptic-scale circulation metrics (Table 3) as a step towards better understanding of its relationship to S_d. As may be expected A and C LWTs showed clear negative and positive correlations respectively with P_s at all sites. Similarly, the N type (mainly positive) and S type (negative) show opposing correlations with $P_{s.}$ The E type showed variable correlations but predominantly negative at summit level and positive at valley level. Both the W and NW types showed mainly positive correlations with P_s, this association being strongest in the far north (Ben More Assynt) for the NW type, but west coast valley locations showed negative correlation for the W type. Results for P_s against the NAO index were variable, dominantly negative for valley locations but with some mountains having a reversed positive relationship at summit level, suggesting that temperatures were low enough at higher levels for the increased precipitation during positive NAO phases to occur as snow.

Finally, original S_d data for the same SSGB locations were evaluated against synoptic-scale circulation metrics (Table 4). All locations showed a negative S_d relationship with the NAO index with the strength of the relationship varying with location. Results suggest that locations that are either more southerly or closer to the coast are more sensitive for S_d with regard to winter NAO variations compared to inland Scottish locations, especially at higher elevations (e.g. Ben Lawers), consistent with Spencer and Essery (2016). However, results do not support the same authors' assertion that S_d at lower elevations is more sensitive to NAO variations than higher altitudes, because altitudinal sensitivity seems to also vary spatially. Analysis based upon LWT frequencies shows further distinctive variations in direction and strength of relationships, which may be referenced against P_s results (Table 3) and relative warmth (S and W types) or coldness (N, E and NW types) of airflow directions (Jones et al. 2014). As may be expected, A and C types showed a clear negative and positive correlation respectively with S_d, and more commonly a stronger relationship at higher elevations. The N type had a clear (positive) correlation only with locations in the north (strongest Ben Macdui and Creag Meagaidh), and at Cader Idris which may be in a similar location regarding increased northerly exposure due to the adjacency of the northern Wales coastline ca.60km away (also indicated by strongly positive correlation of P_s with the N type). By contrast, the E type showed clearer positive correlations with valley-level S_d compared to at higher elevations; a notable exception to this generalisation was a strong correlation at Cross Fell which is more exposed (in terms of adjacency to the coast) to easterly airflows. The S type seemingly mirrors the pattern of the N type by having negative S_d correlations that were strongest at some northerly locations. By contrast, a much clearer pattern is exhibited by W type correlations which showed a clear negative relationship with S_d that is apparently stronger at higher levels, in contradiction to the positive relationship found for P_s. Finally, the NW type shows a positive correlation with S_d that is stronger for northerly locations, similar to relationships described for N type frequency.

4 Discussion

GB snow cover maps presented here show good similarity at lower elevations with previous work (e.g. UKMO map in Kay 2016; Bell al. 2016) and equivalent areas in European-scale remote sensing exercises (Dietz et al. 2012b). However, important map detail is now provided for upland GB areas facilitating interpretation of large-scale altitudinal variations beyond those derived for specific sites (e.g. Trivedi et al. 2007) or from snowmelt modelling (Spencer 2016a). This has allowed provisional estimates of reductions in extent for those areas with larger S_d values (Table 1) where biophysical and socioeconomic implications are likely to be most pronounced. The logistic function demonstrated a strong relationship between S_d and T_m at both upper and lower altitudes, consistent with similar use of this function for interpreting winter snow cover relationships in the European Alps (Hantel et al. 2000; Hantel and Hirtl-Wielke 2007). By including a wide altitudinal range of observations, and hence range of T_m values, the logistic function becomes distinctly non-linear.

Inspection of regression residuals from Figure 2 suggests other influences on S_d beyond the dominant influence of T_m. Residuals can identify seemingly erroneous observations from particular SSGB locations as an inevitable feature of volunteer-derived data. For example, large negative residuals associated with observations of Helvellyn (Cumbrian Fells) from Patterdale station suggest anomalous underestimates of S_d compared to other SSGB locations or more recent observations of nearby Fairfield (Johnson 2005). Differences may also represent unusual topography or local land surface characteristics that favour or disfavour snow cover that are not included in the model. Beyond individual station anomalies, positive or negative residuals showed some clustering in specific years but also with differences based upon whether these were mountain or valley-level observations. There is also an indication of geographical bias in residuals, with positive residuals occurring more commonly in easterly GB mountain locations whereas negative residuals are more common in western mountain locations. This may identify geographical variations in precipitation sources or the influence of oceanicity, as explored further below; or it could be associated with the role of topography in providing larger snow accumulation zones on the more rounded eastern mountains compared to the rugged western mountains, with snow then transferred by prevailing winds to sheltered and shaded NE-facing slopes.

By comparison with T_m, no significant universal relationship with precipitation was found, and various factors may account for this. Firstly, greater spatial variability of precipitation means that the source data (UKMO 5km gridded data) used is generally of lower accuracy, especially in mountain areas due to local orographic effects (Legg 2014). It is also very likely that the original data used for the gridded interpolation are inaccurate because of precipitation undercatch in rain gauges (Pollock et al. 2018), especially during snowfall. Derived estimates of solid precipitation (P_s) were rudimentary but constrained by lack of reliable data, a consequence of difficulties in measuring snowfall in windy oceanic climates with fluctuating temperatures. The precipitation phase equation used here could be further refined in relation to temperature thresholds (e.g. Kienzle 2008), but inaccuracies in measured precipitation from rain gauges and local P_s variations due to topography probably represent the dominant uncertainties. Studies on topographic snowfall effects from other countries have shown considerable variation, even over small distances depending on prevailing wind directions (e.g. Grunewald et al. 2014). In GB these effects are likely to be further modified by the influence of coastal adjacency on moisture supply, as suggested by the site-based correlation analysis (Table 3). Despite these caveats, the present findings are consistent with other European larger-scale studies showing snow cover extent and duration appears primarily dependent on temperature (Scherrer et al. 2004; Henderson and Leathers 2010). Although results suggested that T_m influence on S_d may decrease at higher altitudes in GB, no evidence was found to imply precipitation becomes a more dominant influence at higher altitude as found in the Swiss Alps above ca.1400m, and Norway above ca.400m (Mysterud et al., 2000; Morán-Tejeda et al. 2013). Hence, no evidence was found to suggest recent decadal increases in winter precipitation in GB mountains (Burt and Howden 2013) may have counteracted long-term average declines in snow cover.

The logistic function (Figure 2) exhibits maximum slope between 0 and 3.5°C T_m where for each 1°C change there is a proportional 0.15 change in S_d (i.e. 36.5 days). This steeper segment represents a zone of maximum snow cover sensitivity which translates into a spatial footprint as exemplified by the marked decline in snow duration in areas of northern England for 1980-2010 (Figure 3d). In the current era of rising temperatures, some locations with lower T_m values can therefore be anticipated to cross a threshold into this higher sensitivity zone leading to accelerated long-term S_d decline. This sensitive zone is defined by a combination of latitude/longitude, elevation, and coastal adjacency, such that a general northwards shift can be projected together with a move inland (especially from the west coast) and upslope. Biophysical and socioeconomic implications imply such sensitive locations should become foci for environmental monitoring.

The sigmoid form of the logistic function may also explain previous discrepancies in observations of snow cover with altitude in GB as T_m has a direct altitudinal relationship through lapse rates. Manley (1971) suggested a uniform altitudinal rate of increased snow duration (1 day/15m above 60m) whereas Jackson (1978) suggested exponential increases in duration up to 400 m, above which increase would be linear. Jackson's interpretation is consistent with the upper (non-linear) and middle (roughly linear) T_m regions of the logistic curve, although the transition altitude may actually vary from 400m depending on local T_m profiles and association with latitude and oceanicity influences (Brown 2017); reversion to non-linear form at some higher altitude is also required as maximum S_d is approached. Previous work also suggested an altitude for maximum sensitivity in S_d (e.g. 400±100m in Harrison et al. 2001; 750m in Trivedi et al. 2007) which would be consistent with the zone of maximum sensitivity in the logistic function; as this zone is defined by T_m, altitudinal associations will also vary according to latitude and oceanicity influences. Snow cover observations have also been used to infer greater altitudinal increases in northern compared to southern regions (e.g. 5 days/100m in Wales and central England compared to 15-20 days/100m in Scotland: Harrison 1993; Stirling 1997; Trivedi et al. 2007). Southern regions with higher T_m would correspond with lower S_d rates of change on the right of the logistic function whereas northern regions with lower T_m correspond to steeper S_d rates in the centre of the logistic curve. Similarly, reference to different parts of the logistic function can help explain how altitudinal increases in S_d vary from 20 days/100m in snowy colder years to less than 10 days/100m during milder winters, and that altitudinal S_d increases were linear in colder winters but became non-linear in warm ones (cf. Harrison et al. 2001).

Snow cover also varies on an inter-annual timescale. The present study shows these shorter-term S_d variations can be associated with different frequencies of synoptic-scale atmospheric circulation patterns, with the key distinction being the relative frequency of milder Atlantic westerly air flow compared to prolonged incursion of colder air from north and east. At GB scale (Figure 5), LWT analysis suggests that sensitivity to airflow frequency is generally greater at higher elevations; an exception to this general pattern is shown by the apparent greater influence of the E type at valley level although this may also be an artefact of the absence of mountains in south/east England (i.e. no higher elevation data). The analysis also suggests that sensitivity to airflow direction is primarily related to coastal adjacency, and hence exposure to snow-bearing winds on a yearly basis, and that at higher elevations larger values for yearly S_d are associated with LWT frequencies that are typically colder and wetter compared to influence of mainly colder LWT types at lower altitudes. However, anomalies remain in this general pattern (e.g. Ben More Assynt) that may either be local variations or a consequence of data deficiencies.

Typical conditions for snowfall in GB are when cold arctic or polar-continental air arrives from east or northeast across the North Sea due either to anticyclone formation north of Scotland or when occurring north of cyclonic conditions over southern England or the Low Countries; these synoptic situations characterise cold snowy winters with extended snow duration even at lower levels (e.g. 1978-1979). Snowfall can also occur from cold arctic or polar-maritime air arriving from west to north directions, and the present study suggests that these airflows are particularly influential for longer snow cover durations in the mountains, especially those in northern and western locations exposed to this direction, as occurred notably in years such as 1966-1967 and 2013-2014. This pattern seems to be related to the incursion of air that is both cold and moisture-laden which deposits larger accumulations of snow on exposed mountains, especially from polar-maritime air, which may feature steeper lapse rates than normal (Pepin 2001; Mayes and Wheeler 2013), and when accompanied by frequent fast-moving depressions with only limited incursion of warmer air to melt snow. By contrast, increased frequencies of warmer southerly and westerly airflows are associated with low annual S_d; despite apparent high snowfall rates in mountains exposed to westerly airflows, snow melt is also correspondingly faster.

Variability in S_d values can also be related to the dominant winter NAO mode. A positive NAO index generally indicates a dominant westerly airflow with milder winters and more frequent cyclonic activity but lower snow-cover, whereas a negative NAO index indicates colder conditions with more frequent incursion of polar air and anticyclone formation as associated with longer snow-cover at lower levels. The strongest relationship with the NAO was found in western coastal locations, with a weaker inland relationship consistent with previous work (Spencer and Essery 2016). However, no clear relationship with altitude was found and more complex geographical patterns may exist at higher elevations based upon the combined local influence of temperature and precipitation. For locations or years with similar T_m values, it can be inferred that a higher P_s would imply larger S_d. Nevertheless, the influence of snowfall on S_d does seems to vary more with altitude than does temperature, possibly associated with the slightly reduced influence of T_m at higher altitude suggested by Table 2 although this may also be an artefact of using a generic lapse rate.

Use of synoptic-scale analysis as provided by the NAO index and LWTs therefore provides additional insight into the combined interannual influence of temperature and precipitation in a similar manner to use of joint distribution analysis (cf. Beniston et al. 2011), highlighting the role of cold-wet and warm-dry combinations in producing the snowiest and least snowy annual durations. However, future patterns of interannual/interdecadal climate variability remain rather uncertain, with climate models suggesting a wide range of potential outcomes as the natural variability of the NAO interacts with anthropogenic radiative forcing, affecting both temperature and precipitation variability (Smith et al. 2016; Deser et al. 2017; Wang et al. 2017). This uncertainty constrains future projections of changing snow cover, meaning that occasional years with extended periods of snow cover cannot be excluded despite projected long-term declines. Such unusual conditions may be more marked at higher altitudes, where duration and extent of snow cover appears to have different synoptic-scale influences than at lower level. A further complication is that extreme snowfall events may be less affected by climate change because they occur across a narrower range of temperatures (i.e. closer to 0°C) compared to mean snowfall (O'Gorman 2014).

Synoptic-scale patterns interact with local influences (coastal adjacency and topography) to determine fine-scale spatial variations in snow cover. Hence, inland upland locations with lesser oceanic influence have less interannual variation compared to western coastal locations. Local topography influences exposure to prevailing airflow patterns and their effects on snow accumulation and snow melt. These interactions may explain apparently anomalous local variations: for example, recent observed trends (2002-2015) towards increased snow duration in the western Cairngorms (Andrews et al. 2016) may be attributable to localised exposure to increased frequency of snow-bearing northwesterly and northern airflows.

The empirical statistical approach employed here provides a reduced-complexity method that is most suited for larger-scale analysis, similar to temperature-index methods but without their calibration challenges, and considerably less data intensive than energy-balance models. Scope remains for further research both into large-scale altitudinal variations and at sensitive locations. Methods may be refined to improve understanding of spatiotemporal variability, including geographically-weighted regression for climate data (Brunsdon et al. 2001; Brown 2017). This could also include adjustments in the lapse rate based upon local seasonal and synoptic conditions (Pepin 2001; Holden and Rose 2011). Data sources may be enhanced by new snow cover, snowfall, or snow depth data, especially if available from higher altitudes. Remote sensing sources that can help resolve existing difficulties for cloud-covered or mountainous areas are likely to be particularly useful, including ground sensors that detect diurnal temperature variations associated with snow cover presence (e.g. Zhang 2005). For historic time-series analysis, further potential exists to data mine archives such as the SSGB, potentially using homogenisation techniques to adjust and synthesise incomplete datasets (cf. Caussinus and Mestre 2004; Reeves et al. 2007). Beyond this, the prospect of new volunteer initiatives to record local snow cover variations at national level may be adopted, including for local mountain landmarks, following successful 'citizen-science' applications for other climate-related phenomena (e.g. phenology: Mayer 2010). In GB, this would also allow useful linkages with volunteer-based efforts to record persistence and survival of summer and early autumn snow patches in mountain areas, including integrated analysis of similar anomalous years with extended snow duration related to prevailing weather patterns (e.g. 2013-2014: Cameron et al. 2015). Finally, good prospects remain for further investigation of synoptic-scale analysis, as through exploration of airflow indices and LWTs, both through observed and climate model data (Turnpenny et al. 2002; Jones et al. 2014).

5 Conclusion

Analysis of archived GB snow cover observation data at both low and high altitudes has allowed derivation through logistic regression of a generalised non-linear relationship with temperature. This logistic function was used to define and map long-term average changes in snow cover duration, thereby referencing the extent of recent declines in snow cover and their projection into the future. The non-linear logistic profile also defines regions that are most sensitive to an accelerated loss of snow cover. No significant snow cover relationship with precipitation was found, including snowfall precipitation as distinguished from total precipitation, which may be due to data difficulties or that the influence of precipitation is more variable in relation to location and altitude and its effects more marked in specific years.

Time-series analysis of paired valley and mountain top locations showed that annual snow cover durations were influenced by synoptic-scale circulation patterns, as detected through the winter NAO index and frequency of Lamb weather types. These relationships show an important regional pattern based upon relative exposure to cold and moist snow-bearing airflows. Most notably, analysis suggests that dominant influences for snow cover duration on mountain tops are often different from adjacent valleys, especially the prevalence of polar maritime or arctic air in delivering longer snow durations in mountains that are more exposed to north and westerly influences. Hence, despite the projected long-term average decline in snow cover, the possibility remains for occasional anomalous future snowy years if circulation patterns are favourable. Uncertainties also remain because climate models currently have difficulty in accurately simulating shorter-term changes in synoptic-scale circulation patterns. Further work is therefore recommended, potentially through integration of ground observations, remote sensing, snow simulation models, and climate models, to investigate changing snow cover patterns and influences at both large-scale and site level across the full range of altitudes.

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Table 1. Inferred long-term average changes in GB snow cover area (km^2) for indicative S_d values

S_d	1960-1990	1970-2000	1980-2010	2050s central
50-100 days	10474.2	8391.2	6878.1	525.8
100-150 days	1898.6	1442.6	1121.9	29.5
150-200 days	201.2	137.5	97.1	0.2
>200 days	2.4	1.2	0.7	0.0

Table 2. Pearson correlation coefficients of S_d with temperature and precipitation data for SSGB locations during 1960-1992

Location	T_m	P_s	P_t
Creag Meagaidh (30yrs*) †	-0.48	-0.07	-0.29
Fersit (30yrs) ‡	-0.72	0.49	-0.11
Cross Fell (31yrs) †	-0.67	0.25	-0.18
Alston (31yrs) ‡	-0.75	0.58	-0.04
Cader Idris (28yrs) †	-0.65	0.56	0.21
Dolgellau (28yrs) ‡	-0.71	0.69	-0.08
Ben Vane (28yrs) †	-0.72	0.11	-0.29
Loch Arklet (28yrs) ‡	-0.58	0.45	-0.22
Ben Lawers (32yrs) †	-0.48	0.25	0.01
Ardtalnaig (32yrs) ‡	-0.77	0.65	-0.20
Ben More Assynt (29yrs) †	-0.73	0.01	-0.33
Cassley (29yrs) ‡	-0.74	0.52	-0.02
Ben Macdui (32yrs) †	-0.53	0.34	0.12
Achnagoichan/Aviemore (32yrs) ‡	-0.81	0.69	0.21
Corserine (29yrs) †	-0.67	0.37	-0.11
Forest Lodge (29yrs) ‡	-0.67	0.50	-0.39

* Years of data †Mountain summit ‡Valley station

Table 3. Pearson correlation coefficients of P_s (Oct-May) against principal Lamb weather types (LWT) and NAO index for SSGB locations during 1960-1992 (same total data years as Table 2)

Location	*LWT A*	*LWT C*	*LWT N*	*LWT E*	*LWT S*	*LWT W*	*LWT NW*	*NAO (DJFM)*
Creag Meagaidh †	-0.21	0.01	0.10	-0.52	-0.49	0.73	0.20	0.55
Fersit ‡	-0.33	0.21	0.25	0.29	-0.52	0.45	0.21	0.19
Cross Fell †	-0.63	0.59	0.43	0.04	-0.42	-0.01	0.15	-0.27
Alston ‡	-0.47	0.50	0.39	0.18	-0.40	-0.14	0.10	-0.46
Cader Idris †	-0.49	0.47	0.47	-0.09	-0.44	-0.05	0.32	-0.17
Dolgellau ‡	-0.28	0.36	0.39	0.45	-0.23	-0.40	0.03	-0.58
Ben Vane †	-0.26	0.02	-0.14	-0.23	-0.30	0.58	0.06	0.37
Loch Arklet ‡	-0.62	0.53	0.02	0.31	-0.08	-0.07	0.01	-0.34
Ben Lawers †	-0.25	0.09	-0.18	-0.25	-0.13	0.54	-0.13	0.44
Ardtalnaig ‡	-0.51	0.50	0.07	0.26	-0.15	-0.15	0.03	-0.41
Ben More Assynt †	-0.19	0.03	0.23	-0.45	-0.61	0.53	0.52	0.30
Cassley ‡	-0.48	0.35	0.34	-0.03	-0.54	0.10	0.50	-0.23
Ben Macdui †	-0.47	0.39	0.41	-0.16	-0.51	0.19	0.23	0.00
Achnagoichan/Aviemore ‡	-0.46	0.45	0.49	0.02	-0.51	-0.04	0.23	-0.28
Corserine †	-0.49	0.47	0.28	-0.13	-0.30	-0.03	0.38	-0.22
Forest Lodge ‡	-0.43	0.55	0.25	0.28	-0.30	-0.33	0.21	-0.58

†Mountain summit ‡ Valley station

Table 4. Pearson correlation coefficients of S_d against principal Lamb weather types (LWT) and NAO index for SSGB locations during 1960-1992 (same total data years as Table 2)

Location	*LWT A*	*LWT C*	*LWT N*	*LWT E*	*LWT S*	*LWT W*	*LWT NW*	*NAO (DJFM)*
Creag Meagaidh †	-0.22	0.36	0.40	0.17	-0.37	-0.45	0.55	-0.31
Fersit ‡	-0.35	0.34	0.26	0.07	-0.22	-0.13	0.30	-0.30
Cross Fell †	-0.38	0.37	0.18	0.42	-0.13	-0.49	0.39	-0.48
Alston ‡	-0.35	0.33	0.00	0.47	0.14	-0.39	-0.18	-0.59
Cader Idris †	-0.46	0.58	0.41	0.23	-0.23	-0.42	0.10	-0.44
Dolgellau ‡	-0.06	0.09	0.28	0.51	-0.06	-0.43	0.03	-0.57
Ben Vane †	-0.66	0.64	0.19	0.31	-0.10	-0.37	0.30	-0.53
Loch Arklet ‡	-0.30	0.38	-0.09	0.27	0.07	-0.29	0.09	-0.48
Ben Lawers †	-0.40	0.51	0.17	0.00	-0.21	-0.23	0.21	-0.27
Ardtalnaig ‡	-0.09	0.12	0.01	0.43	-0.02	-0.33	0.04	-0.41
Ben More Assynt †	-0.35	0.34	0.27	0.31	0.05	-0.47	0.25	-0.55
Cassley ‡	-0.29	0.30	0.09	0.41	-0.12	-0.24	-0.03	-0.45
Ben Macdui †	-0.31	0.43	0.36	0.00	0.28	-0.42	0.51	-0.34
Achnagoichan/Aviemore ‡	-0.49	0.40	0.36	0.26	-0.35	-0.27	0.43	-0.42
Corserine †	-0.33	0.37	0.26	0.35	-0.23	-0.34	0.21	-0.62
Forest Lodge ‡	-0.16	0.19	-0.01	0.42	-0.08	-0.38	0.19	-0.52

†Mountain summit ‡ Valley station

1

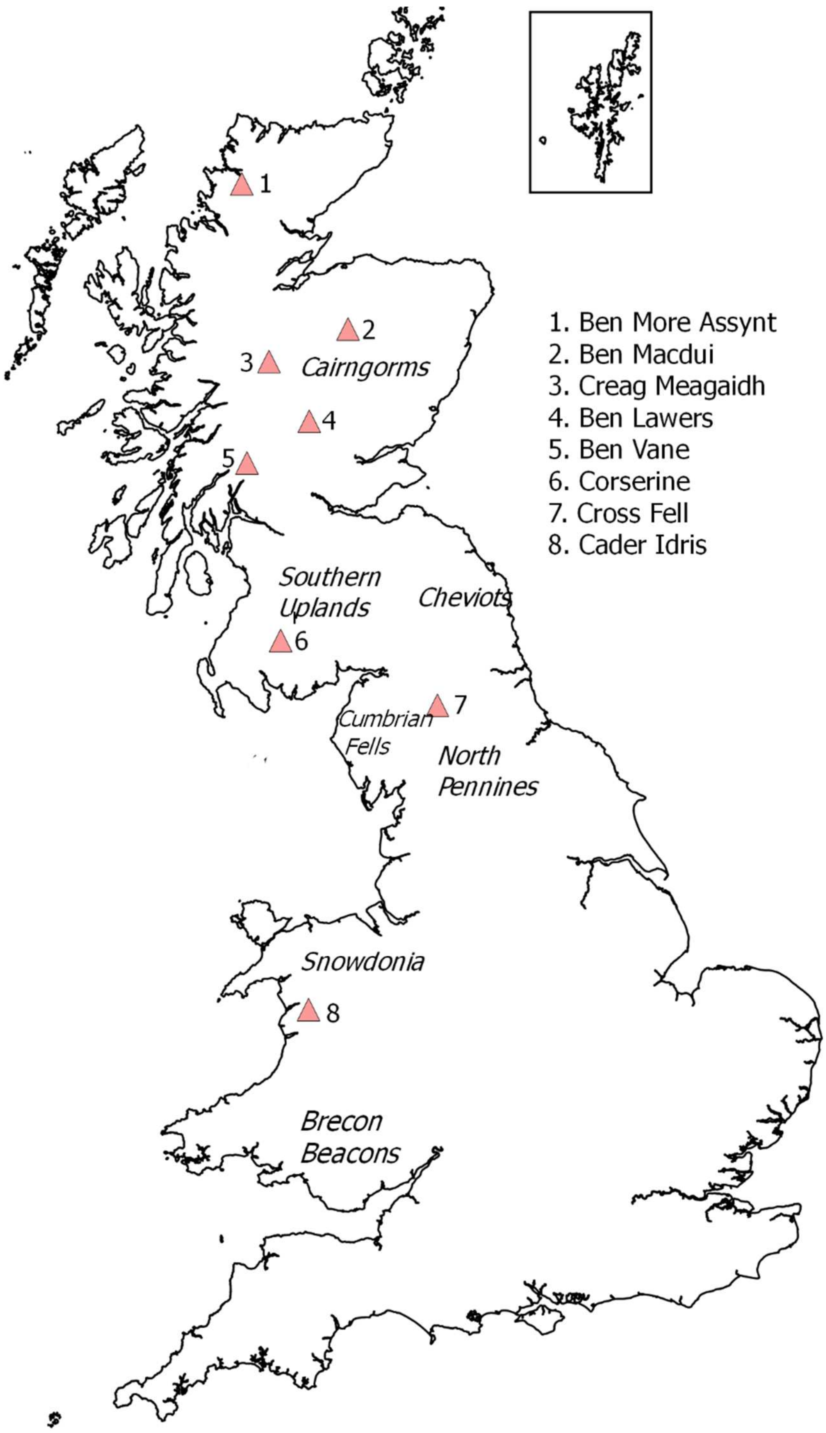

Figure 1. GB locations including mountain peaks used for detailed analysis

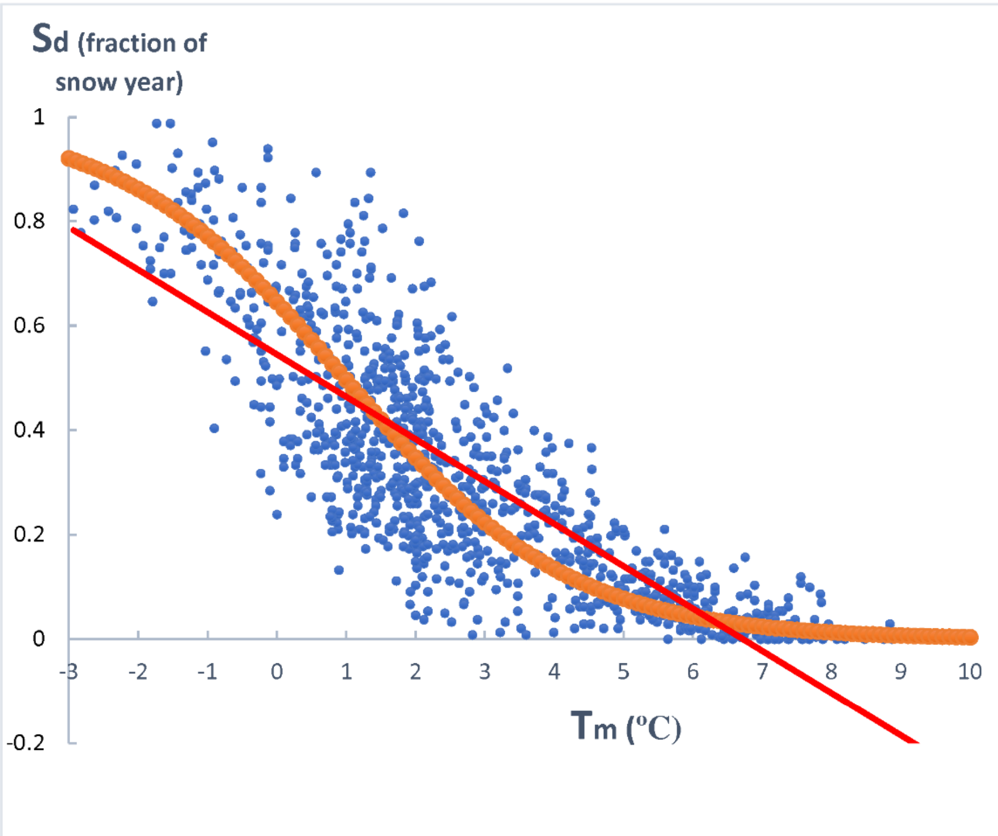

Figure 2. Snow cover duration (Sd) against mean temperature (Tm) for GB, fitted with linear and logistic regression functions

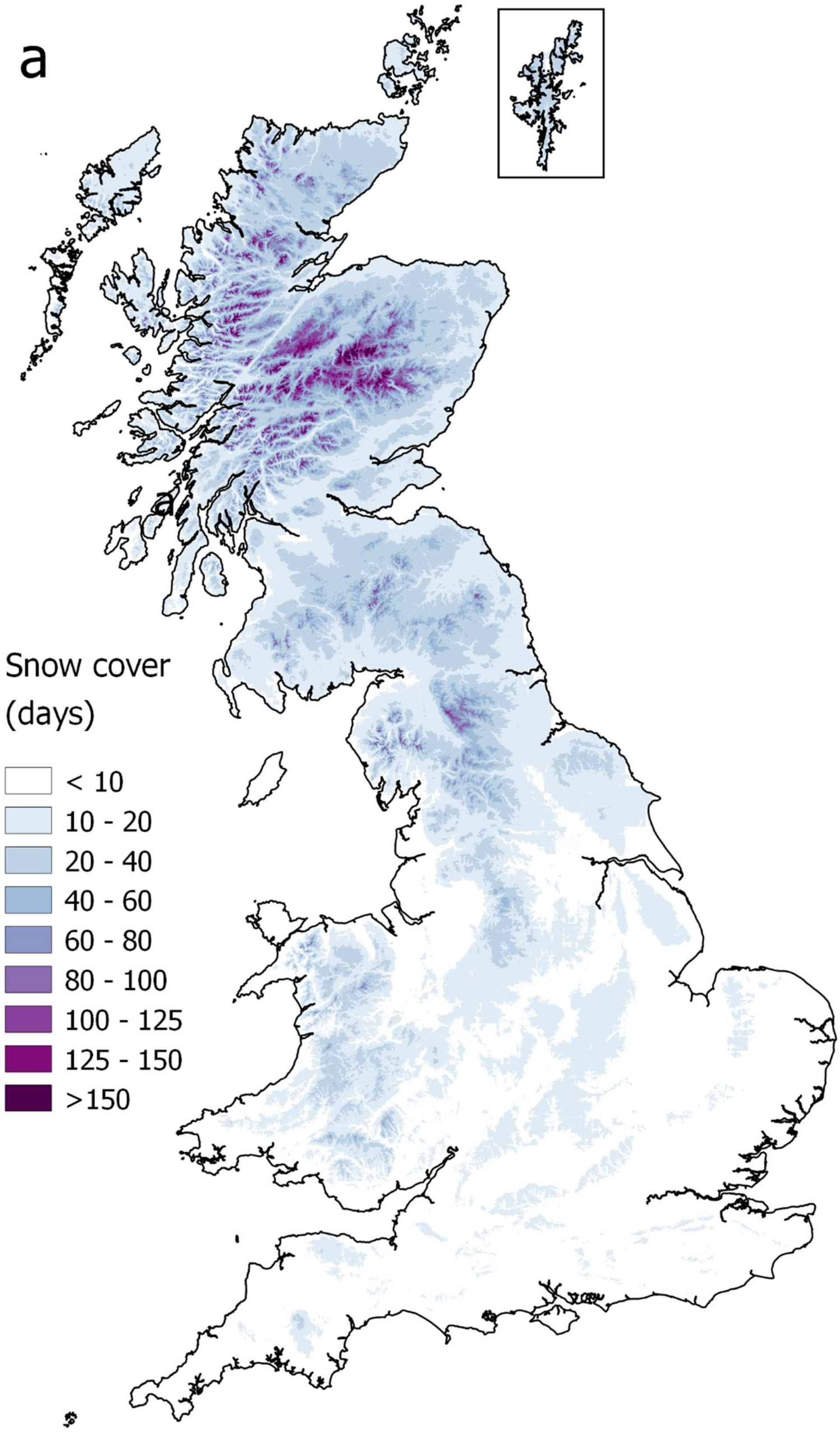

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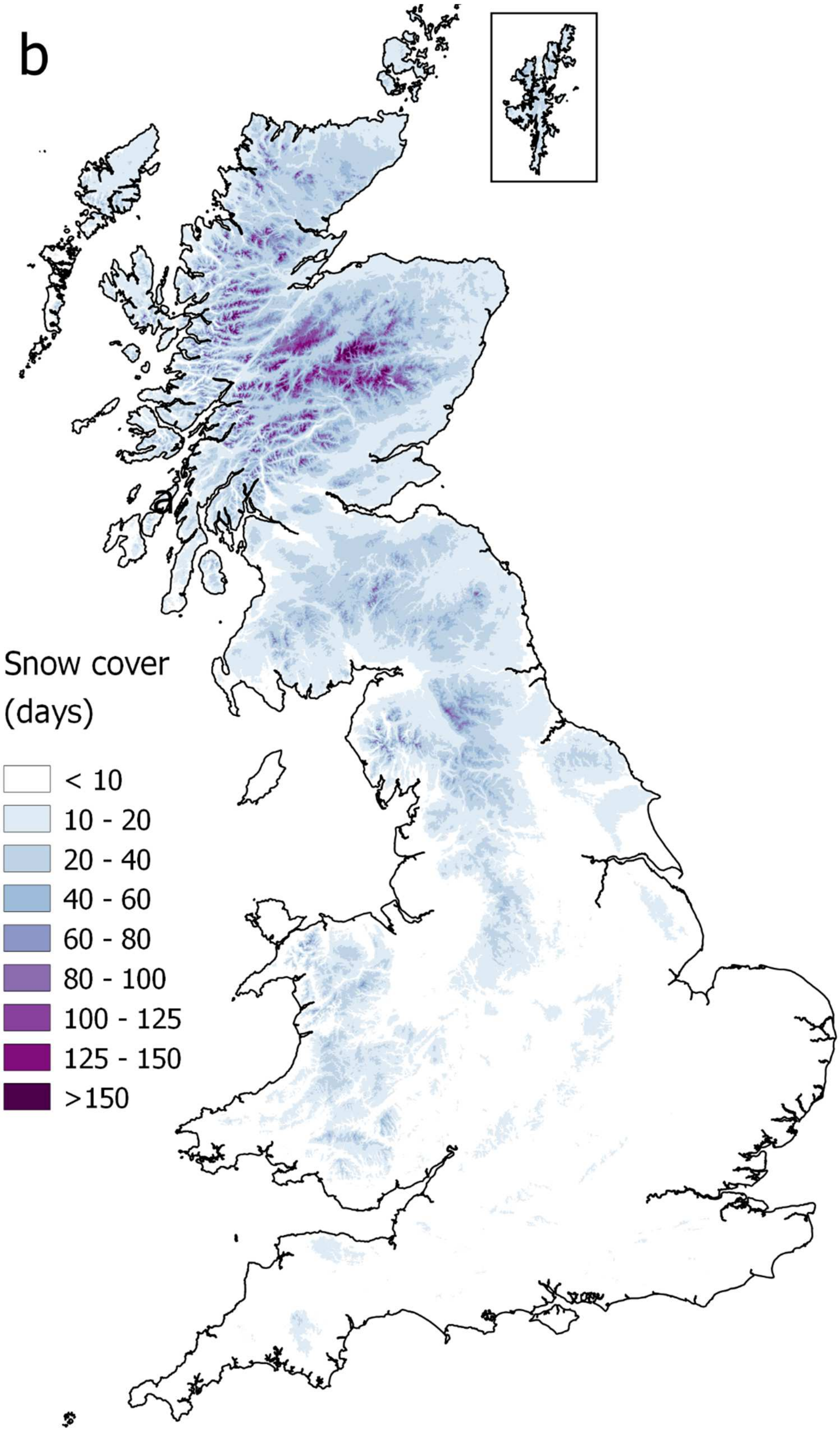
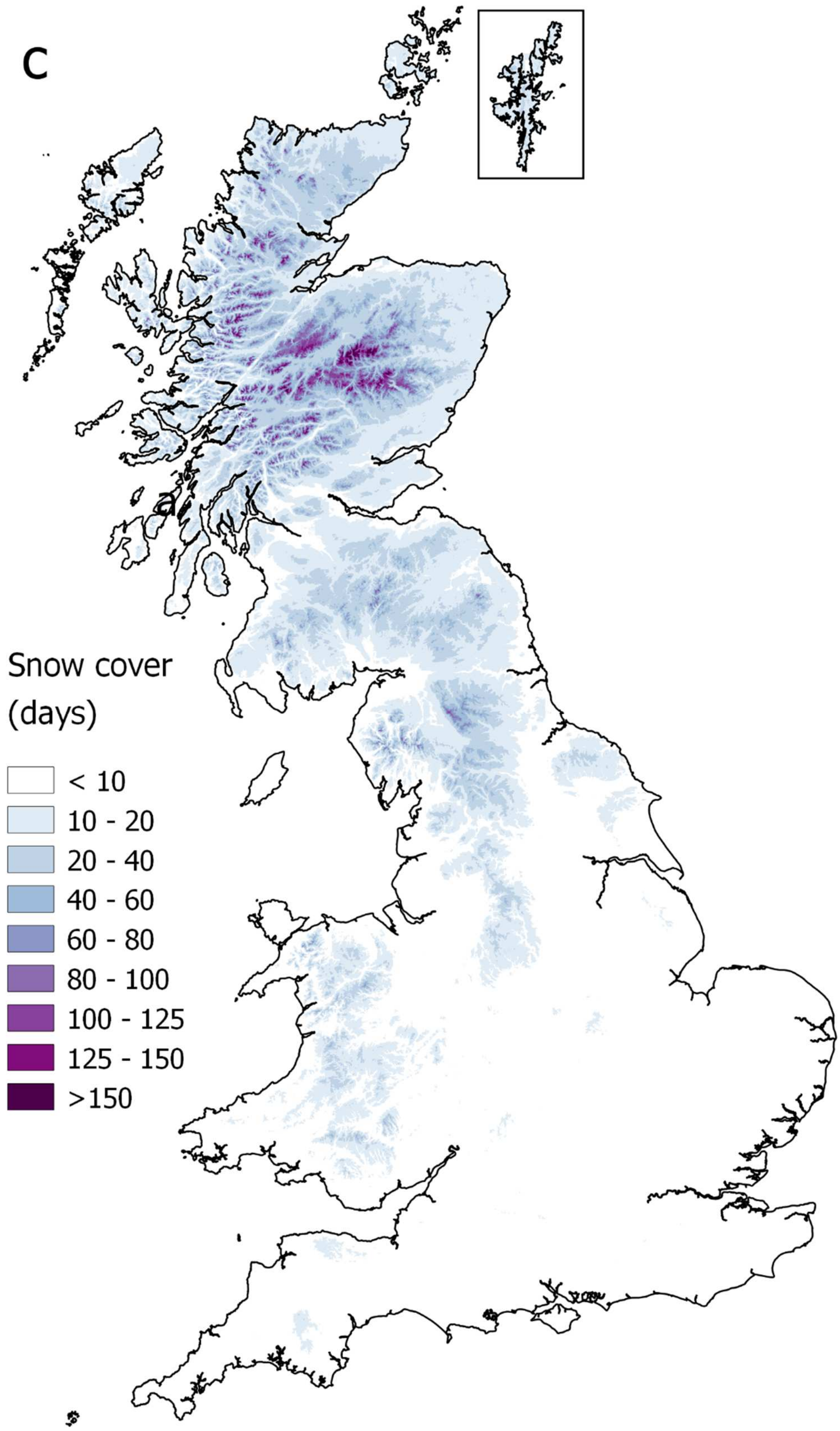
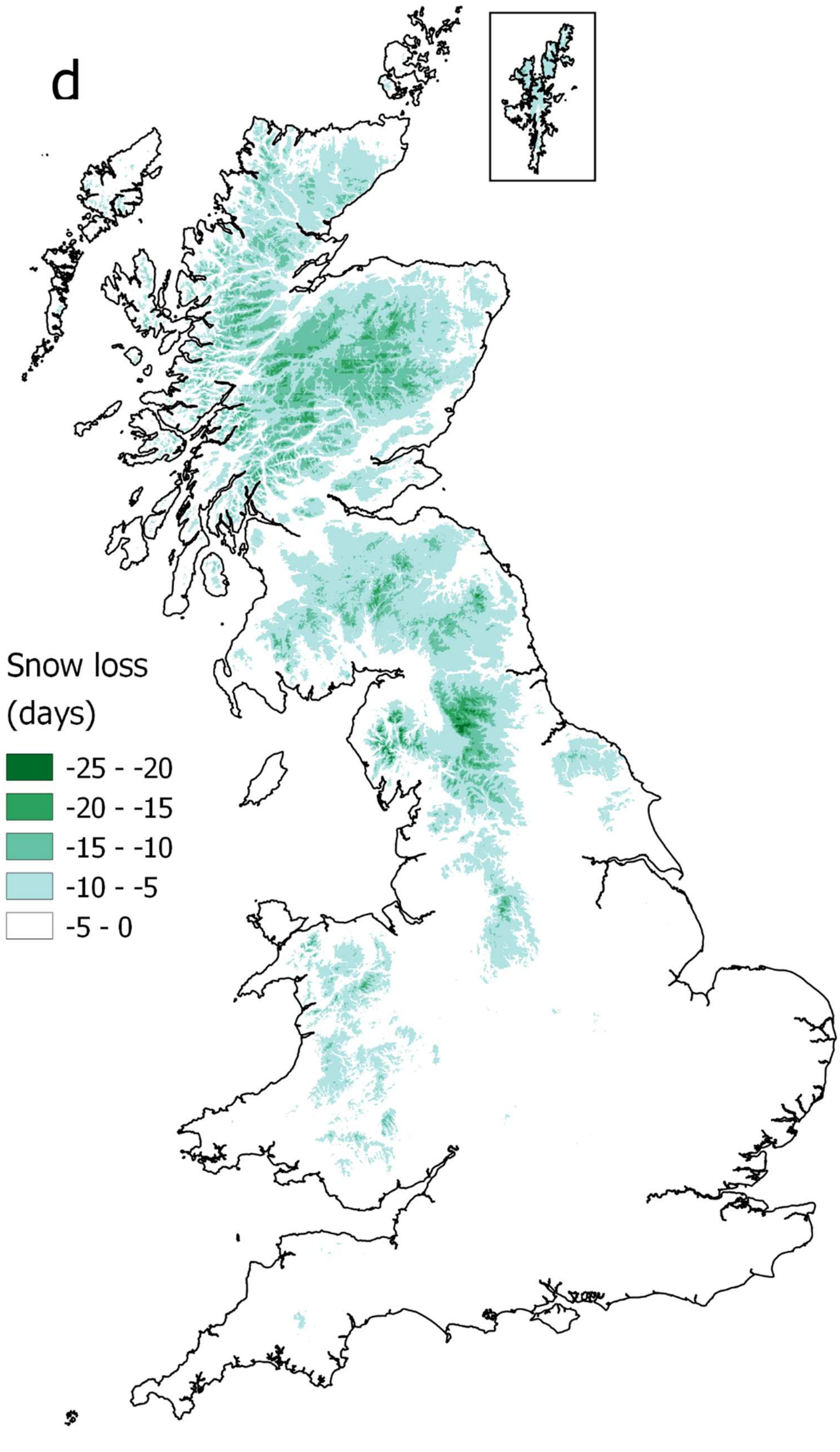
Figure 3. Inferred snow cover duration (a) 1960-1990 (b) 1970-2000 (c) 1980-2010 (d) reduction in 1990-2010 compared to 1960-1990


b
Snow cover
(days)
< 10
10 - 20
20 - 40
40 - 60
60 - 80
80 - 100
100 - 125
125 - 150
>150


c
Snow cover
(days)
< 10
10 - 20
20 - 40
40 - 60
60 - 80
80 - 100
100 - 125
125 - 150
>150


d
Snow loss
(days)
-25 - -20
-20 - -15
-15 - -10
-10 - -5
-5 - 0

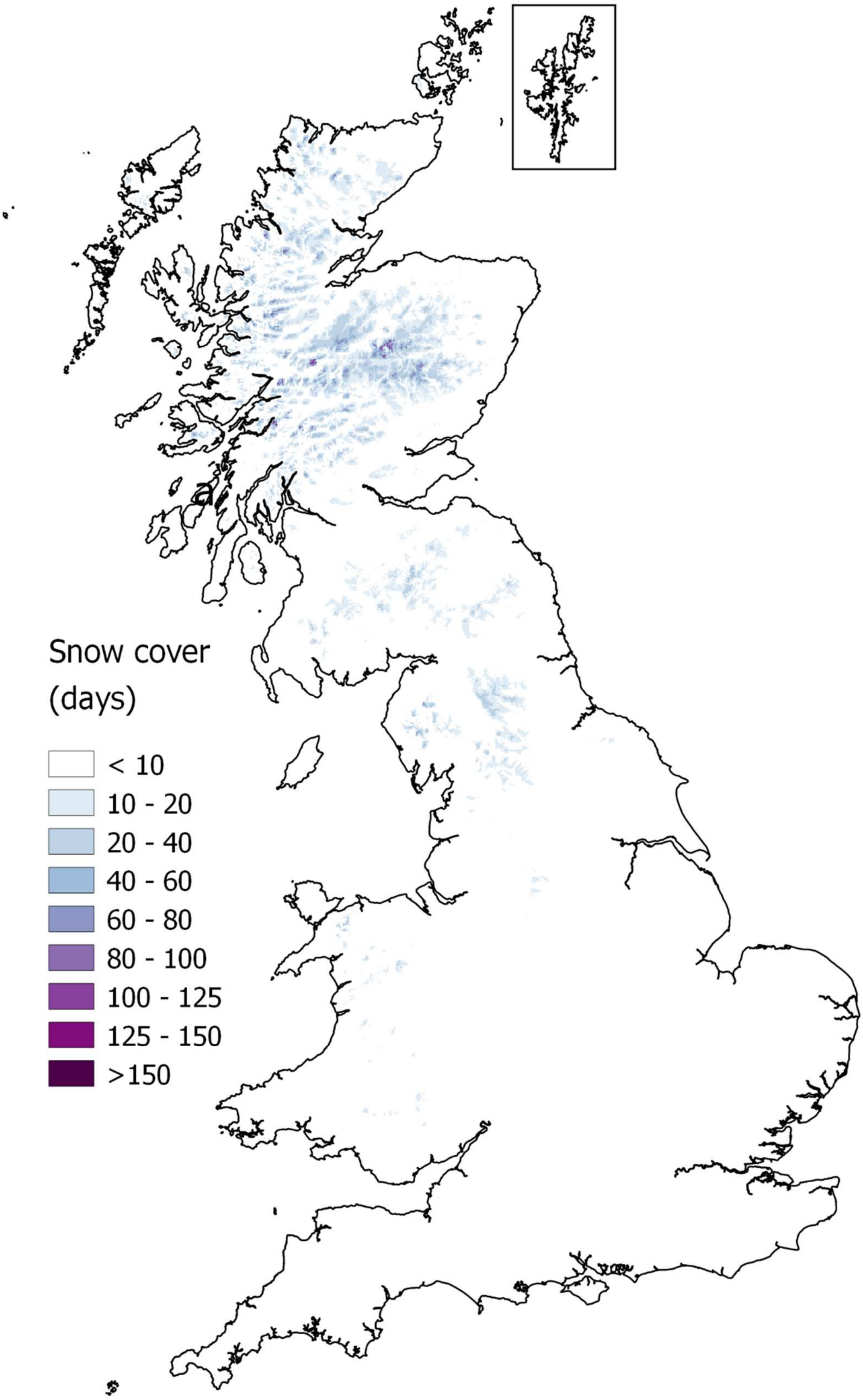


Figure 4. Inferred snow cover duration for 2050s central ensemble projection (A1B emissions)

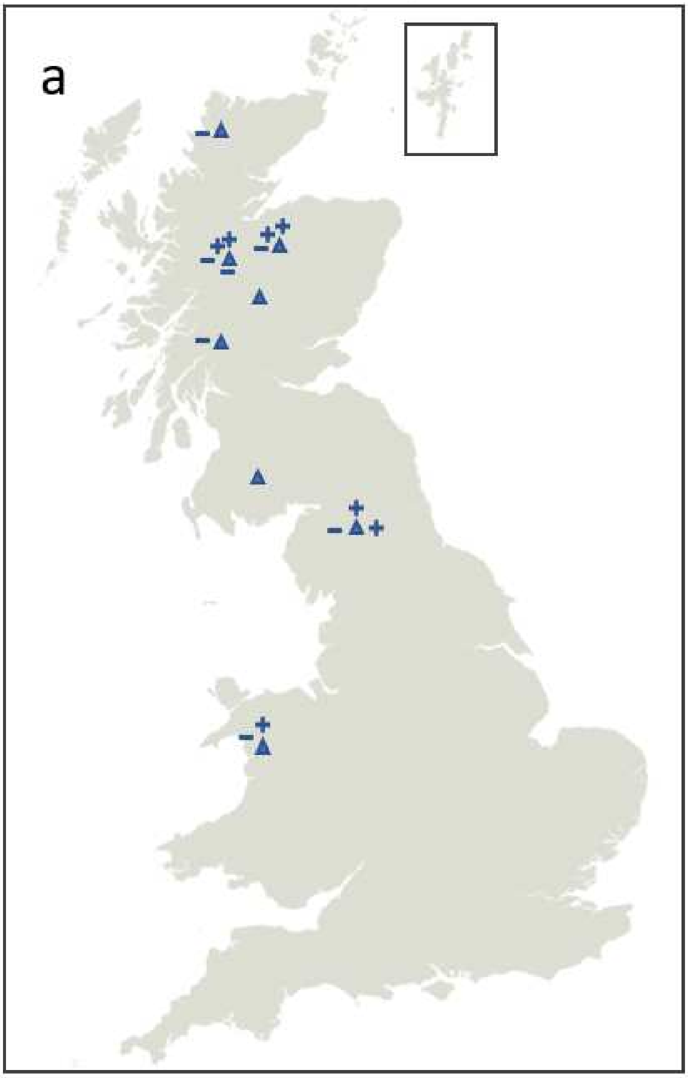
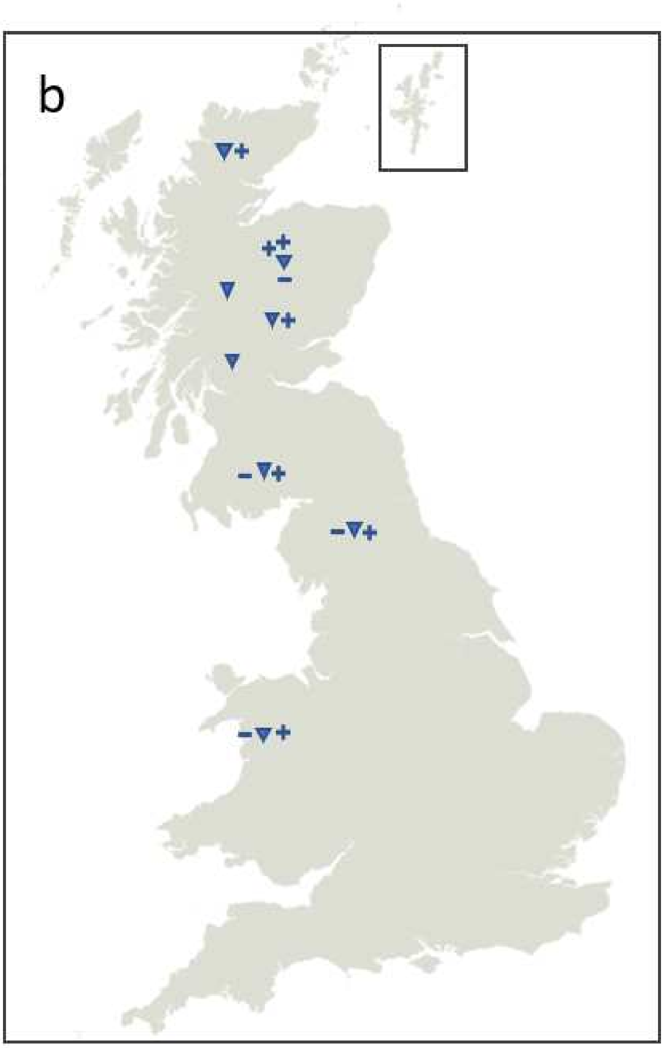


Figure 5. Significant correlations with LWT directional airflow types at reference SSGB sites (a) mountain peaks (b) base stations. Compass position of + (positive correlation) or - (negative correlation) symbol relative to peak Δ or base station ▽ indicates airflow direction of significant relationship.


b